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LUMPED AND DISTRIBUTED PARAMETER MODELS OF THE SELTJARNARNES GEOTHERMAL FIELD, SW-ICELAND

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ABSTRACT

Many mathematical models have been developed for geothermal modelling. Analytical methods are used in lumped parameter models and numerical methods are used to solve the distributed parameter models. For the Seltjarnarnes geothermal field both these methods are used for simulating the fluid flow, chemical mass and heat transport. Data observations from about 30 years, for water-level and chloride concentration are used in the modelling. The main purpose in using a lumped model is to determine two kinds of storages. One is controlled by elastic storage that depends on liquid and formation compressibility in the reservoir and the other by mobility of the free surface of the aquifer. The results are compared with the results of the distributed model.

The distributed parameter model is used for simulation of flow, mass and heat transport in the field. Both leakage and free surface are considered in the model. Future predictions of the behaviour in the field have been calculated from the year 1994 to 2005. Two cases are considered, one with a constant average yearly rate of 35 l/s another with a 2% annual increase from 35 l/s pumping rate. The two cases are considered with and without injection. The injection rate is constant at 20 l/s with the temperature at 50°C and chloride concentration of 500 ppm. The final water level is -55 and -70 m respectively, for the two cases without reinjection, and -5 and -20 m with 20 l/s reinjection. Chloride concentration increases to 2000 ppm in the year 2005 without reinjection. With reinjection, there is little difference between the two cases of pumping rates, but the chloride concentration decreases to 1000 ppm in the year 2005. There is no change in temperature for the future predictions.

1. INTRODUCTION

Geothermal reservoir engineering has been rapidly developing over the last two decades. The main task of the geothermal reservoir engineer is to adjust the parameters and the boundary conditions and predict the future behaviour of the reservoir during production and reinjection. Vatnaskil Consulting Engineers developed the AQUA programme package for use in the modelling of flow, mass and heat transport in geothermal reservoirs.

The Seltjarnarnes geothermal field is located in the town of Seltjarnarnes, which is a suburb of Reykjavik, the capital of Iceland. Since the drilling of the first well in 1965 a considerable amount of data has been collected and several reports describing the field have been published. In 1977 an overview report on the characteristics of most of the geothermal areas within and around Reykjavik was published (Jens Tomasson et al., 1977). A paper about infiltration of seawater into the Seltjarnarnes geothermal field was written in 1986 by Hrefna Kristmannsdottir, and a reservoir studies report of the Seltjarnarnes geothermal field was published in 1987 (Tulinius et al., 1987).

The behaviour of a reservoir in a long term exploitation is controlled by elastic storage that depends on liquid and formation compressibility at the beginning of production. Later in the production history, however, it is controlled by mobility of the free surface. The first one acts in the vicinity of the well and is a short-term behaviour of the reservoir and the other represents the response of the free surface and is a long-term behaviour which is necessary to take into account in geothermal modelling.

In this report the author uses a lumped parameter model and the distributed model AQUA in modelling the Seltjarnarnes geothermal field in order to obtain a better understanding of the response to exploitation of the field.

2. THEORETICAL BASIS

Geothermal reservoir engineering is the study of fluid flow, mass and heat transfer. The purpose of it is to obtain information on properties and physical conditions in a geothermal system, and predict response of the reservoir and wells during exploitation. Many mathematical models have been developed for geothermal modelling. Basically, these can be subdivided into three groups, empirical, analytical and numerical methods. Empirical methods try to fit analytical functions to the available data. A typical example is the decline curve analysis. Analytical methods involve solutions of ordinary or partial differential equations constrained by initial and boundary conditions. This includes the so-called lumped parameter models. Numerical methods are used to solve the distributed parameter models. The following differential equation is the basis of the mathematical model, which can be solved by analytical and numerical methods:

$$a\frac{\partial u}{\partial t}+b_i\frac{\partial u}{\partial x_i}+\frac{\partial}{\partial x_i}(e_{ij}\frac{\partial u}{\partial x_j})+fu+g = 0 \tag{1}$$

The equation is two dimensional, i, j are indicating the x, y coordinates axis.

2.1 Lumped parameter models

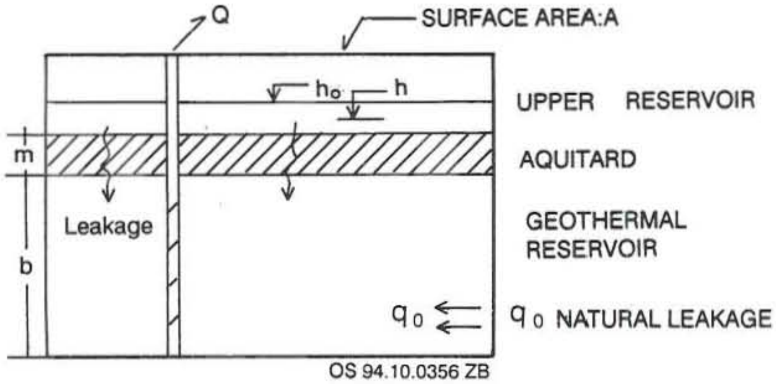

FIGURE 1: Schematic figure of a lumped model

The lumped parameter model is the simplest model to use for geothermal reservoir simulation. Figure 1 shows a schematic figure of a lumped model. Usually it describes the response and behaviour of the geothermal reservoir in terms of only a few parameters. It does not consider the internal distribution of mass and energy. Attention is restricted entirely to the system itself and to what crosses the boundaries as time is the only independent variable. The model is usually solved by ordinary differential equations representing mass and energy conservation. The advantage of this model is its relative simplicity and easy operation. The method does not need very complicated programming and can be solved in a short period. Usually, it is used as a first stage in a modelling process and for checking the results of more complex modelling. The main disadvantage of the lumped model is that it does not consider fluid flow within the reservoir, and does not consider the spatial variations in reservoir properties and conditions (Vatnaskil Consulting Engineers, 1990).

2.1.1 A single phase fluid with leakage

In the lumped model (Figure 1) the continuity equation of flow can be written as

$$Q = \frac{Ak}{m}(h_0 - h) + q_0 - AS\frac{dh}{dt} \tag{2}$$

where

A	= Surface area of geothermal reservoir (m^2);
h	= Pressure head in geothermal aquifer (m);
h_0	= Pressure head in upper layers (m);
k	= Permeability of aquitard (m/s);
m	= Thickness of aquitard (m);
q_0	= Natural recharge (m^3/s);
S	= Storage coefficient of geothermal reservoir.

Furthermore, we define the pressure drawdown s (m), and time constant K (s) as follows:

$$s = h_0 - h \qquad and \qquad K = \frac{mS}{k} \tag{3}$$

Rearranging terms and solving for the drawdown s gives

$$s = \int_0^t \frac{Q(\tau)}{AS} e^{-(t-\tau)/K}\, d\tau \tag{4}$$

2.1.2 A single phase fluid with free surface

The continuity equation for the geothermal reservoir is given by

$$AS\frac{dh}{dt} = Q + \frac{\varphi}{K_f}(h_s - h)\, A_s \tag{5}$$

where A, S, h and Q are defined as above, and

A_s	= Surface area of upper reservoir (m^2);
h_s	= pressure head in upper aquifer (m);
ϕ	= porosity of upper aquifer.

and the time constant K_f is defined by

$$K_f = \frac{\varphi m}{k} \tag{6}$$

The continuity equation for the free surface is given by

$$\frac{dh_s}{dt} = \frac{1}{K_f}(h_s - h) \tag{7}$$

We now define a new time constant and mass withdrawal as follows:

$$K = \frac{K_f}{1 + \dfrac{A_s \varphi}{AS}} \qquad and \qquad M = \int_0^t Q\,dt \tag{8}$$

And finally for the pressure head in the geothermal reservoir we get

$$h = \int_0^t \left(\frac{Q}{AS} + \frac{M}{ASK_f}\right) e^{-(t-\tau)/K}\,d\tau \tag{9}$$

2.2 Distributed parameter models

Distributed parameter models allow a much more detailed description of a reservoir system and the different flow regions that occur in the system. They can be used to simulate the entire geothermal system, including reservoir, caprock, bedrock, shallow cold aquifers, recharge zones, even tectonic structures. In general, it is necessary to use distributed parameter methods for a complete, realistic solution to geothermal problems. Distributed parameter models can be solved by using numerical methods to simulate fluid flow, mass and heat transfer in single phase liquids and two phases as well. It can solve linear and non-linear problems.

2.2.1 AQUA flow model

In transient flow situations, Equation 1 reduces to

$$a\frac{\partial u}{\partial t} + \frac{\partial}{\partial x_i}\left(e_{ij}\frac{\partial u}{\partial x_j}\right) + fu + g = 0 \tag{10}$$

The parameters in Equation 10 are defined as

$$u = h; \qquad e_{ij} = T_{ij}; \qquad f = 0; \qquad g = Q + k/m\,(h_0 - h); \qquad a = -S$$

using x and y as indices instead of i and j, then Equation 10 becomes

$$\frac{\partial}{\partial x}\left(T_{xx}\frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(T_{yy}\frac{\partial h}{\partial y}\right) + \frac{k}{m}(h_0 - h) + Q = S\frac{\partial h}{\partial t} \tag{11}$$

where

h	= Pressure head (m);
h_0	= Pressure head in upper aquifer (m);
k/m	= Leakage coefficient (1/s), where k is the permeability of the semi-permeable layer and m is its thickness;
Q	= Pumping/injection rate (m^3/s);
S	= Storage coefficient.
T_{xx}	= Transmissivity along the principal axis (m^2/s);
T_{yy}	= Transmissivity perpendicular to the principal axis (m^2/s);

In long term exploitation, the storage in the reservoir is controlled by compressibility of the water and the rock in terms of the elastic storage coefficient as in confined aquifers and sometimes by the delayed yield effect. So the complete equation for transient flow is

$$\frac{\partial}{\partial x}(T_{xx}\frac{\partial h}{\partial x}) + \frac{\partial}{\partial y}(T_{yy}\frac{\partial h}{\partial y}) + \frac{k}{m}(h_0 - h) + Q = S\frac{\partial h}{\partial t} + \alpha\varphi\int_0^t \frac{\partial h}{\partial t} e^{-\alpha(t-\tau)}\,d\tau \qquad (12)$$

where

α	= $1/K$ and K is a time constant (s);
φ	= Effective porosity.

For steady state conditions, Equation 1 reduces to

$$\frac{\partial}{\partial x_i}(e_{ij}\frac{\partial u}{\partial x_j}) + fu + g = 0 \qquad (13)$$

The parameters in Equation 13 are defined as

$$u = h; \qquad e_{ij} = T_{ij}; \qquad f = 0; \qquad g = Q + k/m(h_0 - h).$$

Using x and y as indices instead of i and j, Equation 13 becomes:

$$\frac{\partial}{\partial x}(T_{xx}\frac{\partial h}{\partial x}) + \frac{\partial}{\partial y}(T_{yy}\frac{\partial h}{\partial y}) + \frac{k}{m}(h_0 - h) + Q = 0 \qquad (14)$$

The following boundary conditions are allowed:

a) Dirichlet boundary condition; the pressure head, the piezometric head or the potential function is prescribed at the boundary as a function of time;

b) Von Neumann boundary condition; the flow at the boundary is prescribed by defining source nodes (recharge or pumping) at the no flow boundary nodes;

c) Cauchy boundary condition; the flow rate is related to both the normal boundary derivative and the head.

2.2.2 AQUA mass transport model

In the case of transient transport of mass, the parameters of Equation 1 are defined as follows:

$u = c;\qquad a = \varphi b R_d;\qquad b_i = v_i b;\qquad e_{ij} = -\varphi b D_{ij};\qquad f = \varphi b R_d + \gamma + Q;\qquad g = -\gamma c_0 - Q c_w$

By using $x\,y$ instead of the indices $i\,j$, Equation 1 becomes

$$\frac{\partial}{\partial x}\left(\varphi b D_{xx}\frac{\partial c}{\partial x}\right) + \frac{\partial}{\partial y}\left(\varphi b D_{yy}\frac{\partial c}{\partial y}\right) - v_x b\frac{\partial c}{\partial x} - v_y b\frac{\partial c}{\partial y} = \varphi b R_d \frac{\partial c}{\partial t} + \varphi b R_d \lambda c - (c_0 - c)\gamma - Q(c_w - c) \qquad (15)$$

where

b	= Aquifer thickness (m);
c	= Solute concentration (kg/m^3);
c_0	= Solute concentration of vertical inflow (kg/m^3);
c_w	= Solute concentration of injected water (kg/m^3);
Q	= Pumping rate (m^3/s);
$v_x v_y$	= Velocity vector taken from the solution of the flow problem (m^3/s);
γ	= Infiltration rate or leakage rate (m/s);
λ	= Exponential decay constant (1/s).

The dispersion coefficients D_{xx} and D_{yy} are defined by

$$\varphi D_{xx} = a_L v^n + D_m \varphi \qquad and \qquad \varphi D_{yy} = a_T v^n + D_m \varphi \qquad (16)$$

and the retardation coefficient R_d is defined by

$$R_d = 1 + \beta(1-\varphi)\rho_s/(\varphi\rho_1) \qquad and \qquad \beta = K_d \rho_1 \qquad (17)$$

where

a_L	= Longitudinal dispersivity (m);
a_T	= Transversal dispersivity (m);
D_m	= Molecular diffusivity (m^2/s);
K_d	= Distribution coefficient;
v	= Velocity (m/s);
φ	= Effective porosity ;
ρ_1	= Density of the liquid (kg/m^3);
ρ_s	= Density of the porous medium (kg/m^3);

2.2.3 AQUA heat transport model

AQUA can solve single phase heat transport problems. The parameters in Equation 1 must then be defined as follows:

$u = T;\qquad a = \varphi b R_h;\qquad b_i = v_i b;\qquad e_{ij} = -bK_{ij};\qquad f = \gamma + Q;\qquad g = -\gamma T_0 - Q T_w.$

By using $x\,y$ instead of the indices $i\,j$, Equation 1 becomes

$$\frac{\partial}{\partial x}(bK_{xx}\frac{\partial T}{\partial x}) + \frac{\partial}{\partial y}(bK_{yy}\frac{\partial T}{\partial y}) - v_x b\frac{\partial c}{\partial x} - v_y b\frac{\partial T}{\partial y} = \varphi b R_h \frac{\partial T}{\partial t} - (T_0 - T)\gamma - Q(T_w - T) \qquad (18)$$

where

T = Temperature (°C);
T_0 = Temperatuer of vertical inflow (°C);
T_w = Temperature of injection water (°C);

The dispersion coefficients K_{xx} , K_{yy} are defined by

$$K_{xx} = a_L v^n + D_h\varphi \qquad and \qquad K_{yy} = a_T v^n + D_h\varphi \qquad (19)$$

and the retardation coefficient R_h by

$$R_h = 1 + \beta_h(1-\varphi)\rho_s/(\varphi\rho_1) \qquad where \qquad \beta_h = C_s/C_l \qquad (20)$$

and

C_1 = Specific heat capacity of the liquid (kJ/kg°C);
C_s = Specific heat capacity of the porous medium (kJ/kg°C);
D_h = Heat diffusivity (m^2/s);

3. THE SELTJARNARNES GEOTHERMAL FIELD

3.1 Location

The Seltjarnarnes geothermal field is located within the town Seltjarnarnes, which is a suburb of Reykjavik, the capital of Iceland (Figure 2). There are several geothermal fields within and around Reykjavik. These fields are typical low-temperature geothermal fields, where the subsurface temperature is below 150°C at 1 km depth. The heat source of these fields is the regional geothermal gradient.

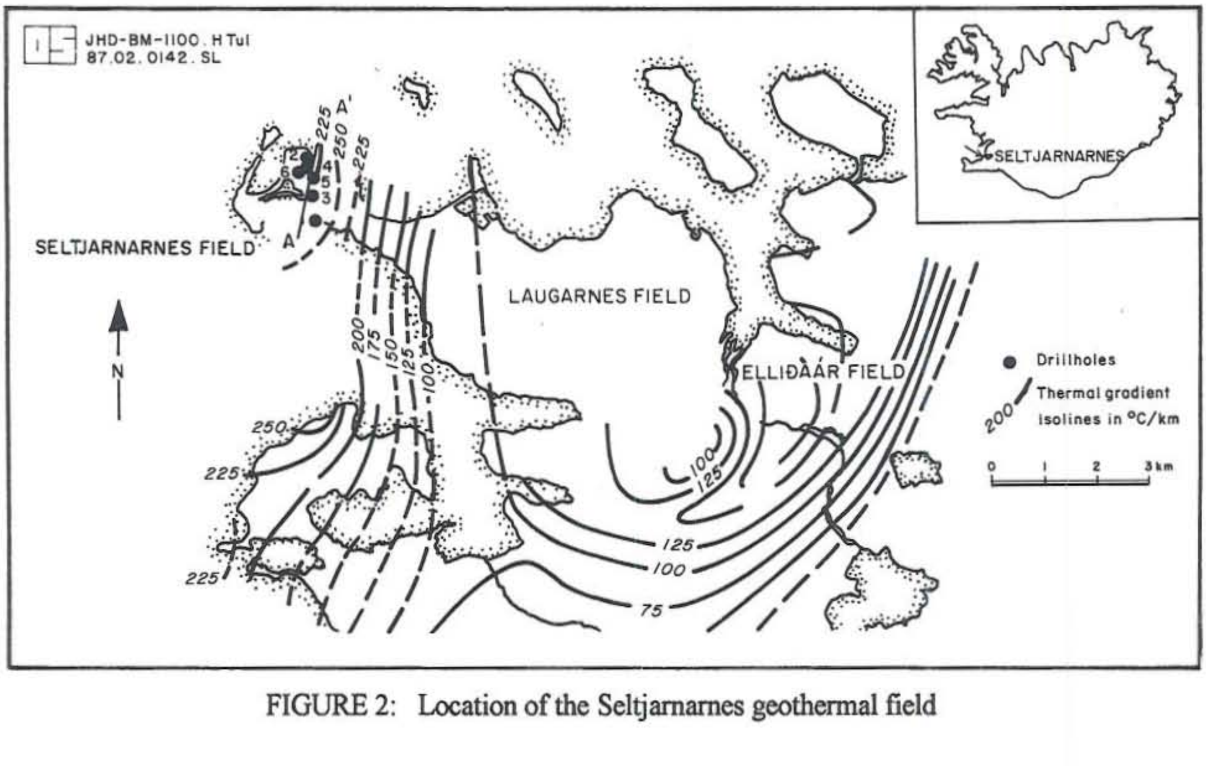

FIGURE 2: Location of the Seltjarnarnes geothermal field

3.2 Geology

The Seltjarnarnes geothermal field and the other geothermal fields around Reykjavik are situated 8-10 km north of the volcanically active Reykjanes rift zone. Figure 3 shows a simplified geological cross-section of the Seltjarnarnes geothermal field, extending from well Sn-01 in the south to well Sn-02 in the north (line A-A′ on Figure 2). The geological profile is based on analyses of drill cuttings collected at 2 m intervals during drilling. The reservoir rocks are mainly 1.8-2.8 m.y. old Quaternary rocks (Tulinius el at. 1987). As shown in Figure 3 the rocks have been divided into seven main groups. Sediments are from surface to 100 m. From 100-400 m basalts and hyaloclastites are found. The bottom of the basalts and sediments is at 600-900 m depth, in the north it is at about 800 m, in the south at nearly 900 m, but in the centre it is at 600 m. Under this layer hyaloclastite is found, at 900-1000 m depth. Tholeiite basalt is found between 1000 and 1700 m, and below 1700 m there is olivine basalt and altered and fresh basalt.

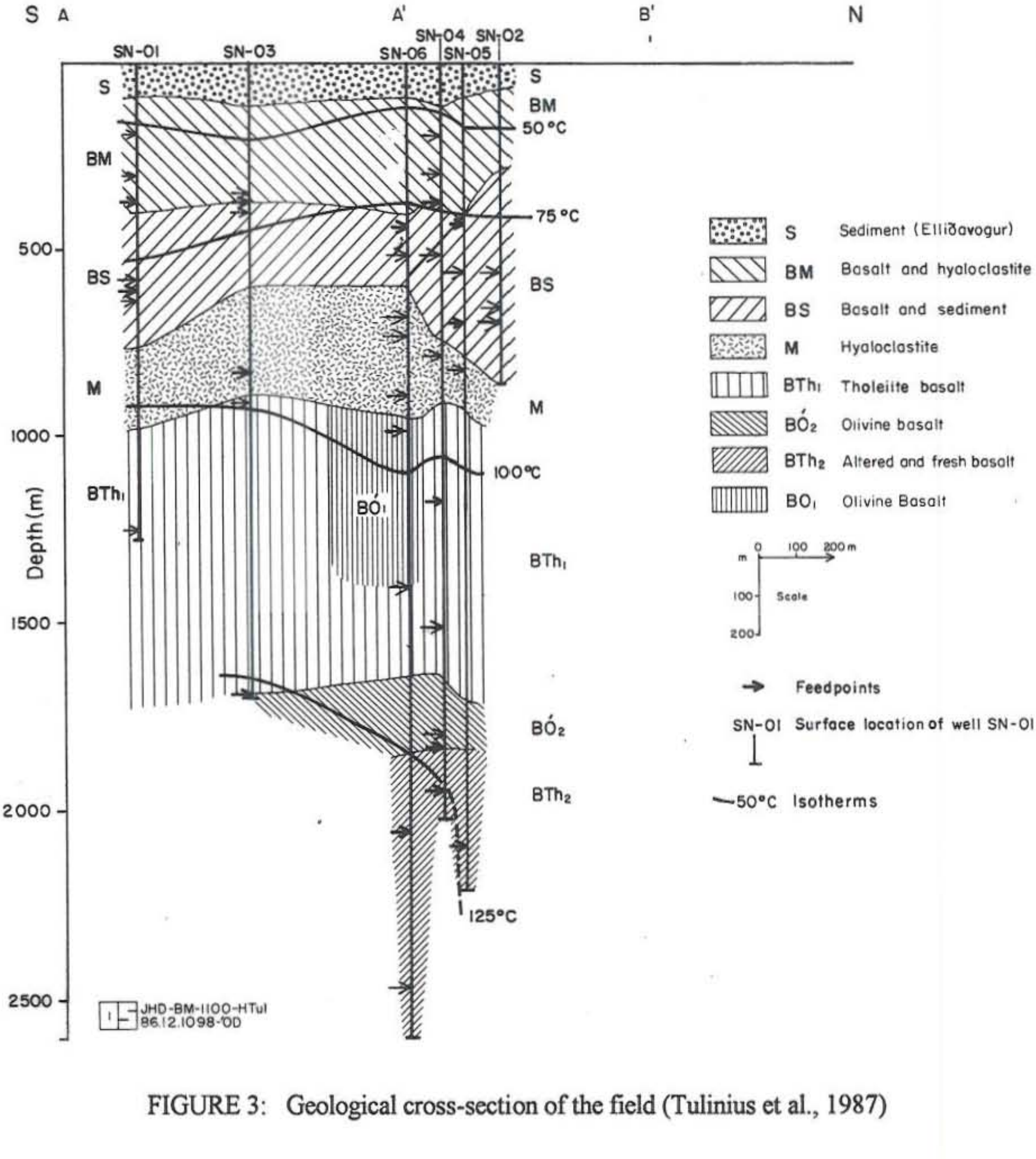

FIGURE 3: Geological cross-section of the field (Tulinius et al., 1987)

3.3 Hydrogeology

The geothermal water of the Seltjarnarnes field is of meteoric origin. Monitoring of the water level in the area has shown that the production from the Laugarnes and Ellidaar fields in Reykjavik has no effect on the water level in the Seltjarnarnes geothermal field. It is, therefore, concluded that the Seltjarnarnes field is separated from the other fields by low permeability rocks.

Based on the information obtained during drilling, several feedzones have been detected in all of the wells. The locations of the feedzones are based on water losses or gains during drilling or from temperature logs during the production period. All known feedzones are shown in Figure 4 as arrows; and the main ones are shown in Figure 3. Table 1 gives the depth, temperature, chloride concentration and a rough estimate of production from each major feedzone. The geothermal field can be divided into three parts in vertical direction. The uppermost is from 400-600 m with a production rate of less than 30% and temperature of 70°C. The middle one is at 1700 m depth, the number of feedzones is less than in the upper one, and the temperature is 100°C. The

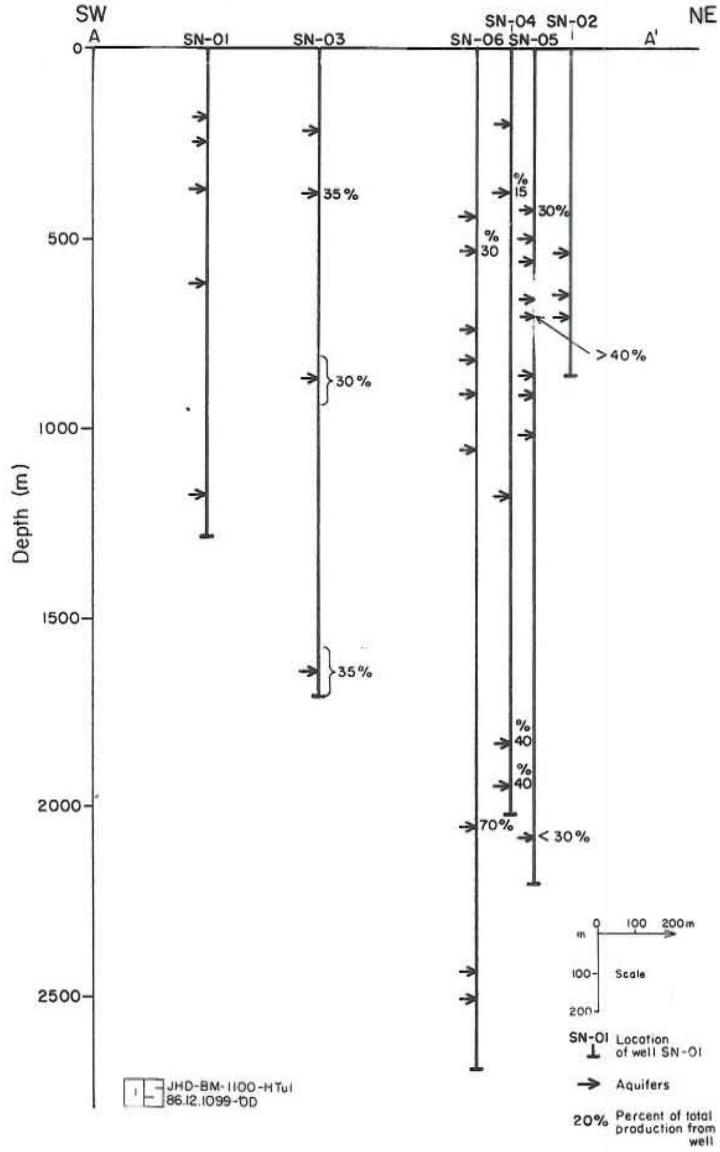

FIGURE 4: Location of feedzones

TABLE 1: Main feedzones in wells in the Seltjarnarnes area (Cl from 1984) (Tulinius et al., 1987)

Well	Depth (m)	Temperature (°C)	Cl concentration* (ppm)	Production (%)
Sn-03	380	72	2100	35
	870-930	100	1000	30
	1700	127	1000	35
Sn-04	370	75	2000	15
	1180	104	900	5
	1840	122	900	40
	1950	126	900	40
Sn-05	430	75	1800	30
	700	90	800	>40
	2090	>120	800	<30
Sn-06	440-700	80	2100	30
	2060	133	1000	70

deepest is below 1700 m in well Sn-06 down to 2700 m. The production from this part is more than 35% of the total production in well Sn-06, and up to 70-80% in Sn-04, with a temperature of 130°C which is the main reservoir in the field. Chemical samples have been collected since 1966. In well Sn-01 and Sn-02 a total of five samples were taken from 1966 to 1969. The chloride concentration for all those samples was similar or around 500 ppm. In 1970, two months after the completion of well Sn-03, the chloride concentrations of the samples taken were 500 ppm at 400 and 1000 m depth. It therefore appears that the initial chloride concentration was rather uniform over the entire field around 500 ppm (Tulinius et al., 1987). But, since 1981, there is a rapid chloride concentration breakthrough. Figure 5 shows the change of chloride concentration with time. It is apparent that the chloride concentration in the well has increased from 500 ppm in 1966 to almost 2000 ppm in 1994, probably because of seawater intrusion. The lower aquifers have the same problem because of the recharge of seawater.

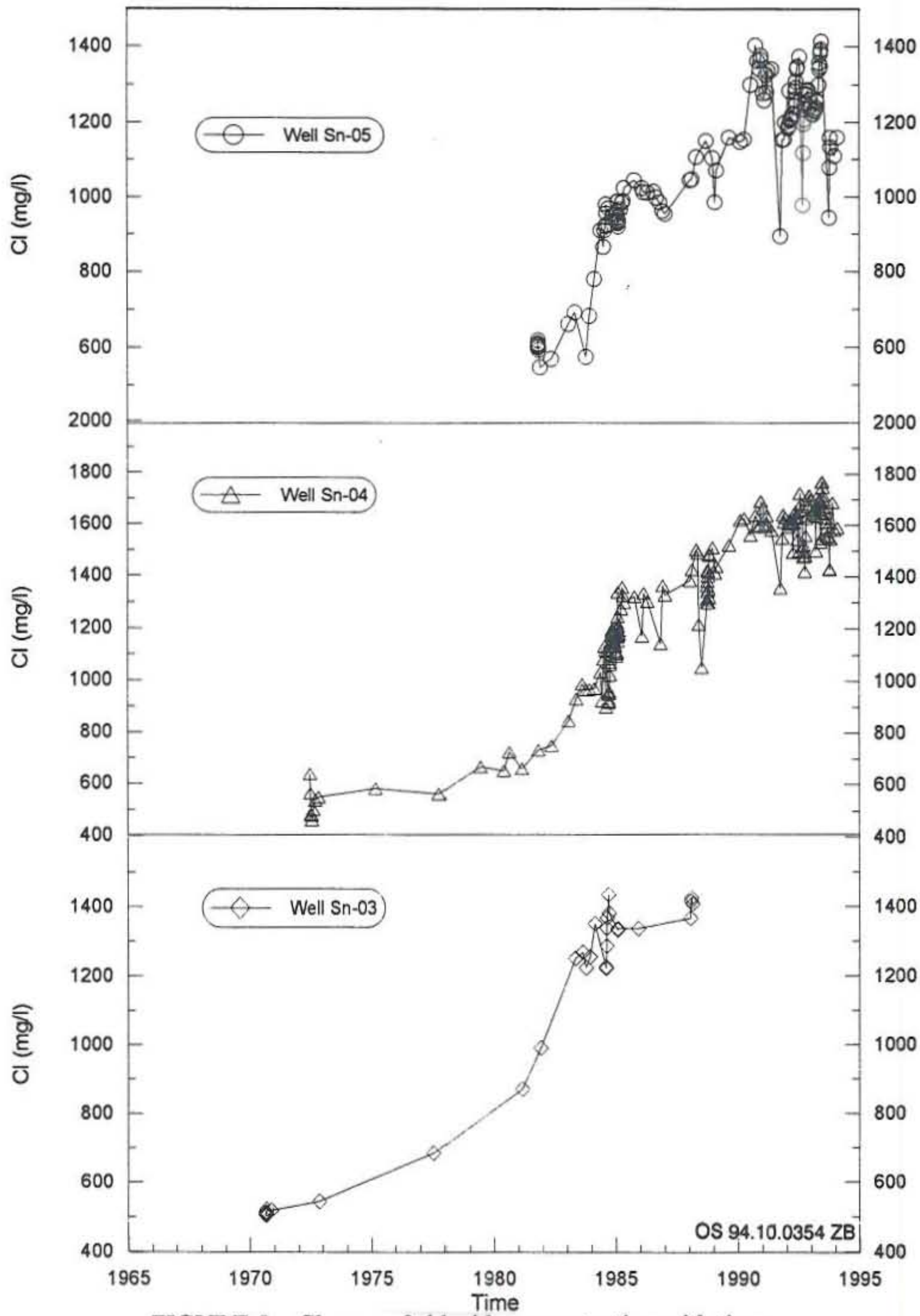

FIGURE 5: Change of chloride concentration with time

Temperature increases with depth down to the bottom of all wells, with about 300°C/km gradient in the uppermost 200 m and about 35°C/km below 600 m depth. The highest temperature is in well Sn-06 at 2700 m depth, more than 140°C. The average temperature for the interval 400-2200 m is about 110°C, where most of the aquifers are located.

3.4 Production history

In the Seltjarnarnes geothermal field, drilling started in 1965. The wells Sn-01 and Sn-02 were originally drilled to estimate the thermal gradient in the area, but were later deepened to be used for space heating in 1969. Well Sn-02 produced about 3 l/s from July 1966 to September 1971. After well Sn-03 was drilled, pumping from wells Sn-01 and Sn-02 was stopped, and they used as observation wells. After the completion of well Sn-04, the Seltjarnarnes Municipal District Heating Service was founded, being supplied with water from Sn-03 and Sn-04. The pumping rate was 40 l/s. The next well Sn-05 was drilled in 1981 to a depth of 2207 m, and well Sn-06 was drilled in 1985 to a depth of 2701 m. Maximum yield of these four production wells is 110 l/s, which is sufficient for space heating of the entire town (Tulinius et al., 1987). The locations of the wells are shown in Figure 2. Table 2 shows the characteristics of the wells. The most productive aquifer in the field is located below 1700 m. In wells Sn-03 and Sn-05 30-40% of the water comes from this aquifer, and up to 80% in wells Sn-04 and Sn-06. The water level was estimated to be 75 m a.s.l. at the beginning of production. But it declines with time because the pumping rate has increased continuously. In 1990, it was at -140 m a.s.l. The demand decreased when flow-meters where placed in every house instead of a system based on a maximum flow restriction. The water level started to recover and is now, in 1994, at -110 m a.s.l. The production history is shown in Figure 6.

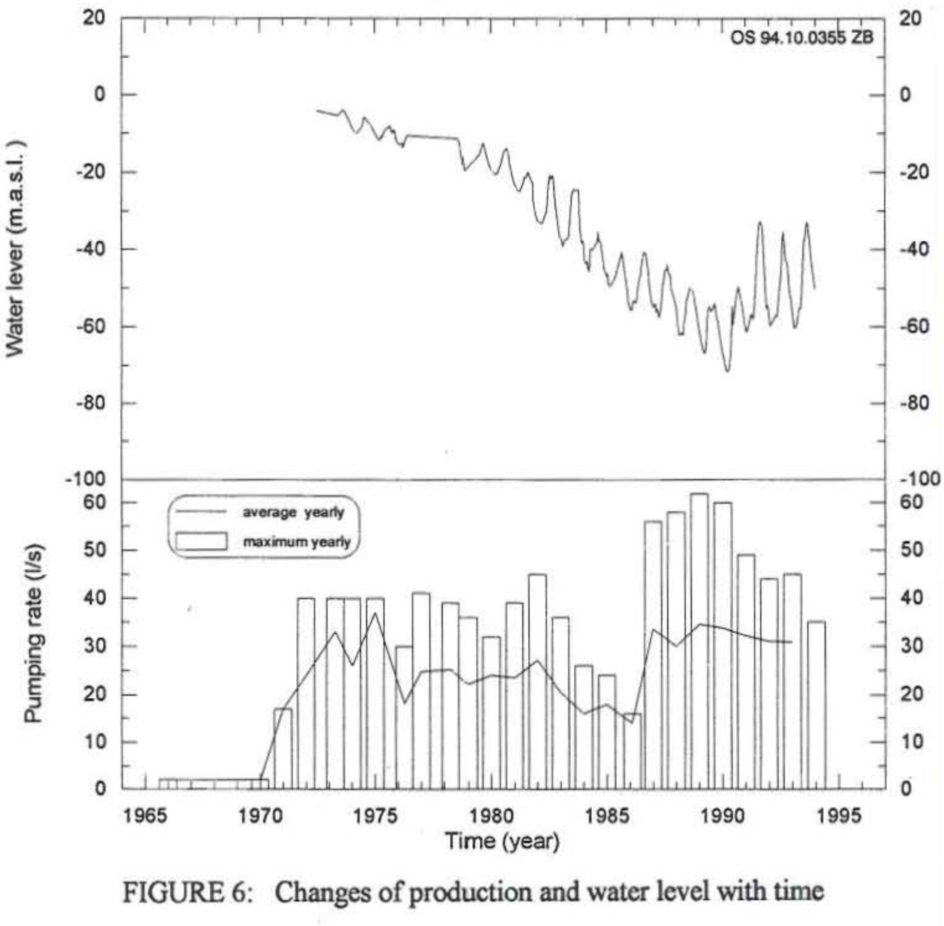

FIGURE 6: Changes of production and water level with time

TABLE 2 Characteristics of the wells in the Seltjarnarnes area (Tulinius et al., 1987)

Well no.	Drilled (year)	Depth (m)	Casing (m)	Temperature (°C)	Max. yield (l/s)
Sn-01	1967	1282	18.5		1-3
Sn-02	1965	856	81.5		3
Sn-03	1970	1715	99	101-103	15
Sn-04	1972	2025	172	111-116	35
Sn-05	1981	2207	168	90-100	30
Sn-06	1985	2701	414	115-117	30

4. MODELS OF THE GEOTHERMAL FIELD

4.1 Conceptual model of the Seltjarnarnes geothermal field

After analysing all the available information on the Seltjarnarnes geothermal field, the conceptual model of the field can be determined. The geothermal water comes from meteoric precipitation falling southeast of the area. The water percolates down, probably to about 3-4 km depth. It is heated up due to the anomalous thermal conditions in the bedrock, and discharged in the Seltjarnarnes field through fissures which extend to considerable depth and are more permeable than in the surrounding area. Due to seawater intrusion, the chloride concentration has increased rapidly since 1981. The Seltjarnarnes geothermal field is separated from other fields by low permeability rocks.

4.2 Lumped parameter models of the Seltjarnarnes geothermal field

Two kinds of lumped parameter models (see Figure 1) are used to fit the water level drawdown of the reservoir. One is a single phase fluid with leakage and the other is a single phase fluid with free surface. The measured drawdown of the reservoir is taken from observation well Sn-02. It has been used for observation for more than 20 years. The production rate is taken from the years 1966-1994.

4.2.1 Calibration of the models

The initial parameters are according to the well test data. First the model with leakage was tried, the parameters giving the best fit show that the storage coefficient needs to be nearly as big as the porosity. The calculated drawdown corresponds to the average value of measured water level (Figure 7) and the leakage is much higher than mass calculations based on the changes of chloride concentration. This means that the reservoir response over a long period not only depends on the elastic storage coefficient but also on the porosity at the free surface.

The free surface model gives more realistic results. The measured and calculated drawdown of the water level and the calibrated parameters are shown in Figure 8. As seen there the storage coefficient is 1.5×10^{-5}, the time constant is 5000 days, the porosity is 0.01, the area of the model is 150 km^2 and the area of the free surface is 40 km^2. The calculated drawdown gives a good fit with the measured one. The result of fitting shows that the free surface is not effective immediately upon the decline of the water table in the main aquifer. The delay time depends on the permeability and the thickness of the semi-permeable layer.

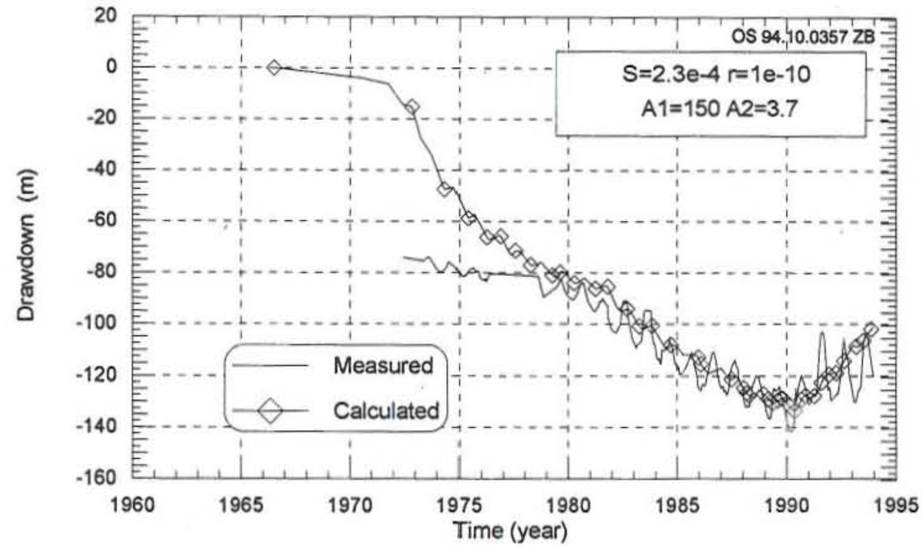

FIGURE 7: Measured and calculated drawdown with leakage in well Sn-02

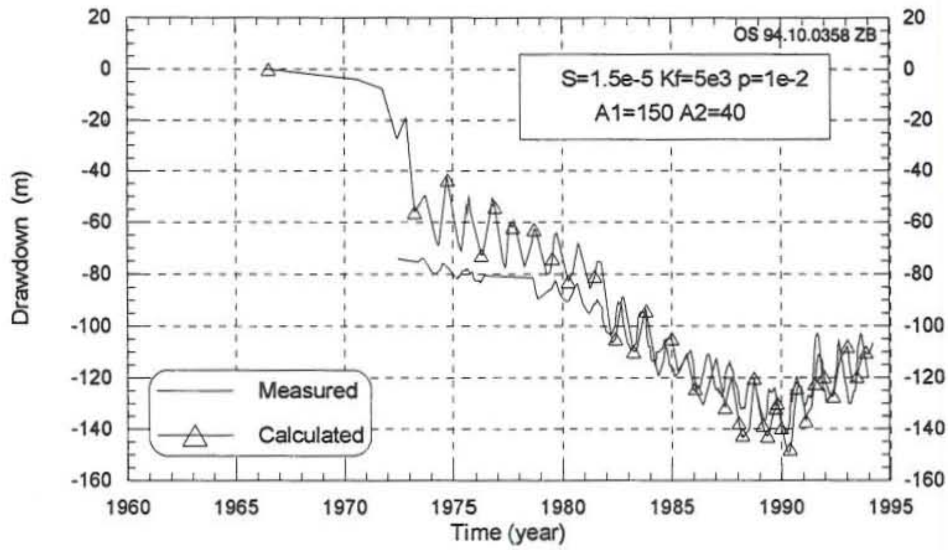

FIGURE 8: Measured and calculated drawdown with free surface in well Sn-02

4.2.2 Model prediction

Future predictions of the water level have been calculated from 1994 to 2005 for two cases using the free surface model (see Table 3):

Case 1: The average yearly pumping rates are 35 l/s. During January to March the pumping rate is 43 l/s, 27 l/s during April to September; and 35 l/s during October to December.

Case 2: The pumping rates are based on case 1 and an annual increase of 2%.

TABLE 3: Pumping rate for prediction

Period of prediction 1995 - 2005	Case 1: constant rate (l/s)	Case 2: Same as Case 1 + annual increase
January-March	43	2%
April-September	27	2%
October-December	35	2%

The results of the calculations are shown in Figure 9. In case 1, the drawdown of the water level will be 30 m from 1994 to 2005, with the water level at -75 m a.s.l. in 2005. In case 2, the drawdown of the water level will be 45 m, i.e. the water level will be at -90 m a.s.l. in the year 2005.

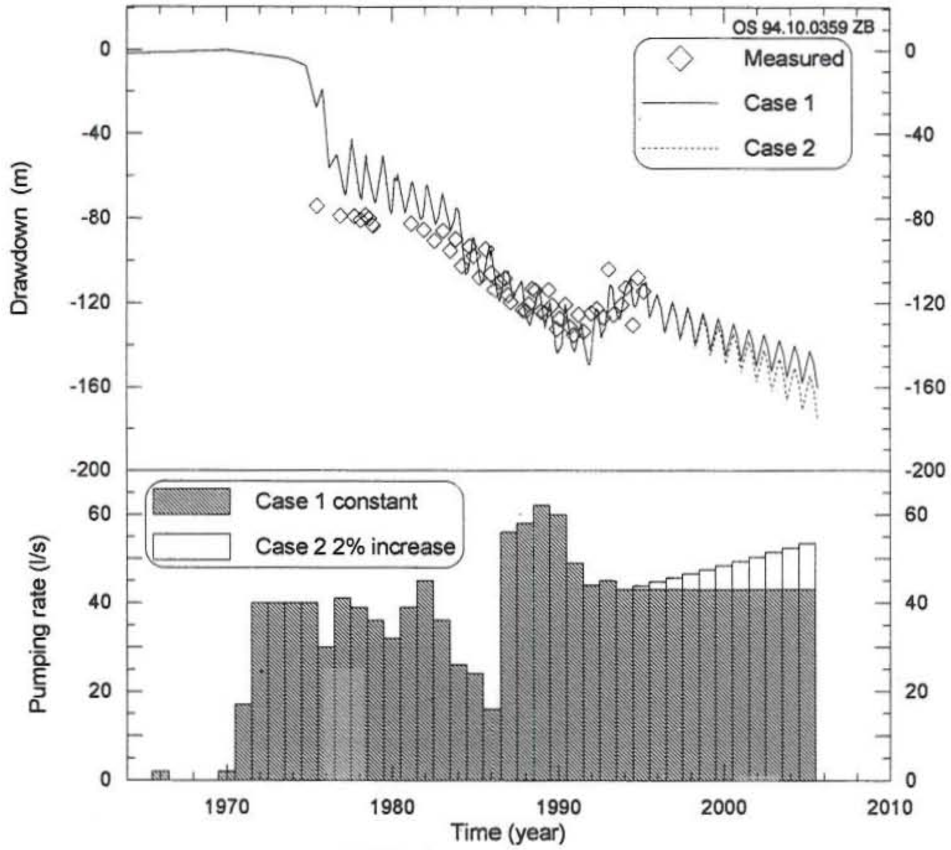

FIGURE 9: Prediction of water level in the Seltjarnarnes field

4.3 Distributed parameter model of the Seltjarnarnes geothermal field

The model area is 150 km^2. The pumping area is located in the middle of the modelling area. An impermeable boundary is given around the whole model area except in the southeastern part, where there is constant head boundary. The mathematical equation is solved numerically by using the finite element method in the AQUA programme. There are 768 elements and 1472 nodes in the finite element mesh, and the mesh is more dense around the pumping area than away from it. Figure 10 shows parts of the nodes in the model area and the constant head boundary nodes.

4.3.1 Flow model

In the AQUA flow model Equation 1 is approximated by the Galerkin finite element method using triangular elements. A constant initial water level was assumed in the whole modelling area before production. The initial value of the water level is 75 m a.s.l. also taken as the constant head boundary condition.

Initial guess for the parameters transmissivity and storativity in the field are taken from the results of well tests (Table 4).

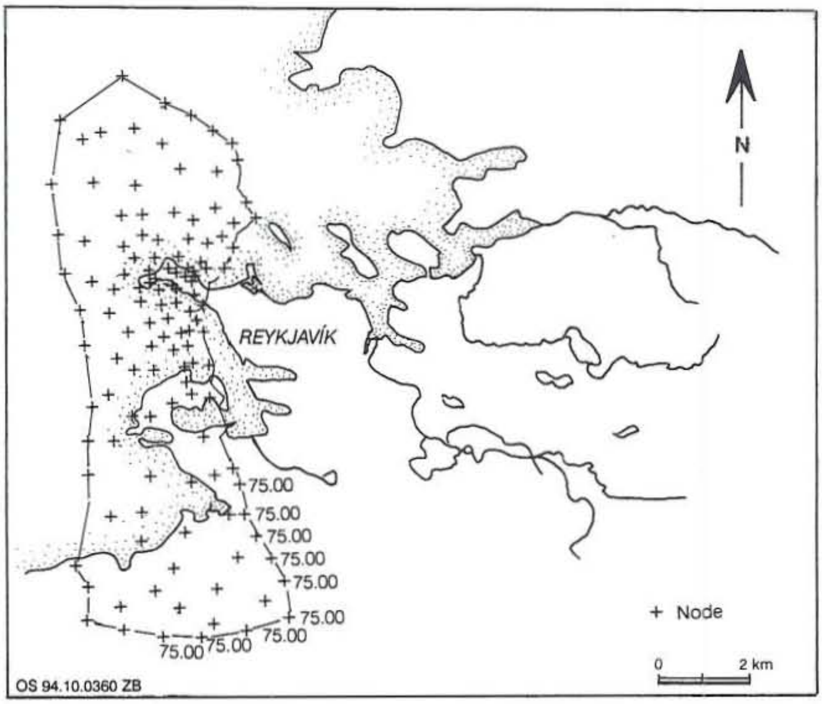

FIGURE 10: The model area for the Seltjarnarnes field

TABLE 4: Well tests in Seltjarnarnes wells (Tomasson et al.1977; Tulinius et al., 1987)

Wells used for pumping	Monitoring wells	Depth (m)	T (m^3/Pa-s)	S (m/Pa)	k (mD)	Date
1	2	500-800	23.4×10^{-8}	4×10^{-9}	33	June 1969
2	2	500-800	13.8×10^{-8}		19	1966-68
3	3	100-1700	$13\text{-}28\times10^{-8}$		29	Mar. 1970
3	1,2	150-1250	11.7×10^{-8}	$1\text{-}2\times10^{-9}$	16	Sep. 1970
3	3	150-1700	$28\text{-}57\times10^{-8}$		60	1970
4	1,2,3	0-1000	11.7×10^{-8}		16	Aug. 1972
4	1,2,3	1000-1700	31.2×10^{-8}		44	1972
2,3,4	2		13.3×10^{-8}		19	1970-80
5	5	170-2200	13.2×10^{-8}	1.9×10^{-9}	19	May 1981
6	6	414-2540	3.13×10^{-8}	2.2×10^{-9}	4	Dec. 1984

The parameters of the model are determined by the method of trial and error, fitting the observed and calculated water level in observation wells Sn-01 and Sn-02. The parameters that gave the best fit are as follows: Transmissivity is 8×10^{-3} - 7×10^{-4} m^2/s; anisotropy, sqrt(T_{yy}/T_{xx}), is 0.3; direction of high permeability is N-S; storage coefficient is 2.1×10^{-4} - 7×10^{-5}; leakage coefficient is 9×10^{-11} - 8×10^{-11} 1/s in the pumping area and there is no leakage outside the area; delayed yield porosity is 3×10^{-3} and the time-constant for delayed yield is 5000 days. Areal distribution of the parameters is shown in Figures 11, 12 and 13. The results of the calibration are shown in Figures 14 and 15. Both leakage and free surface effects are considered in the model.

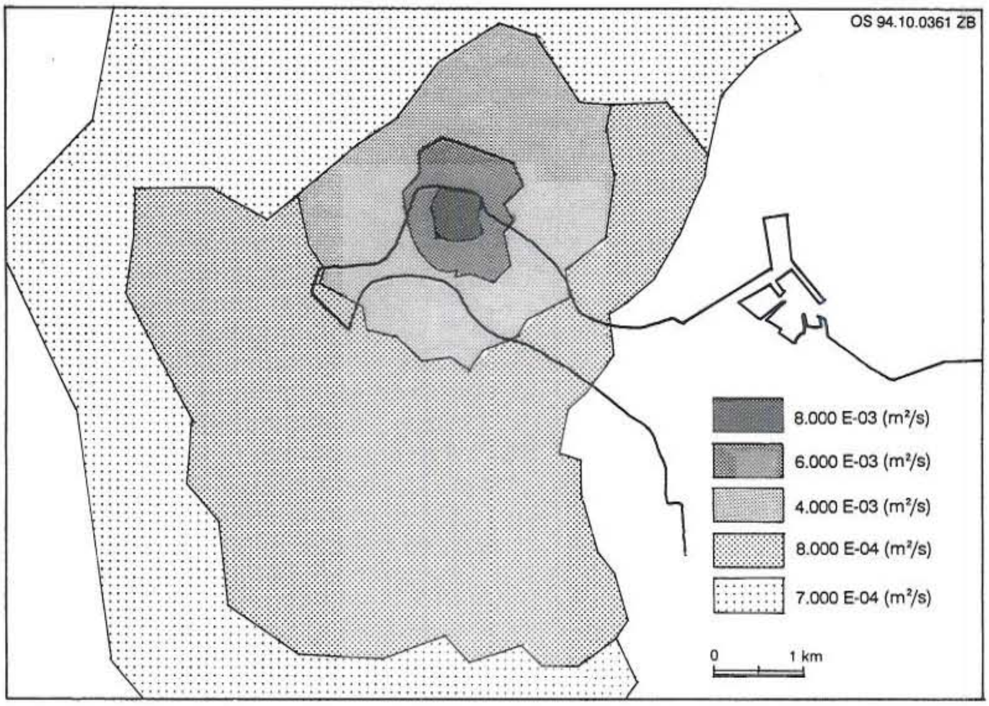

FIGURE 11: Transmissivity distribution for the Seltjarnarnes field

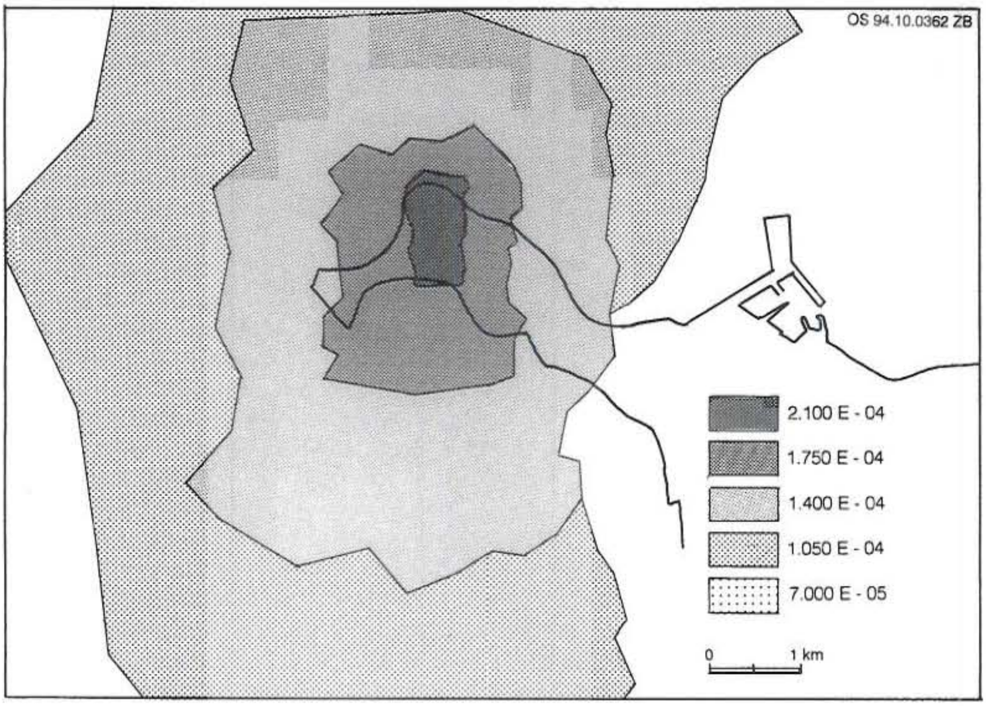

FIGURE 12: Storativity distribution in the Seltjarnarnes field

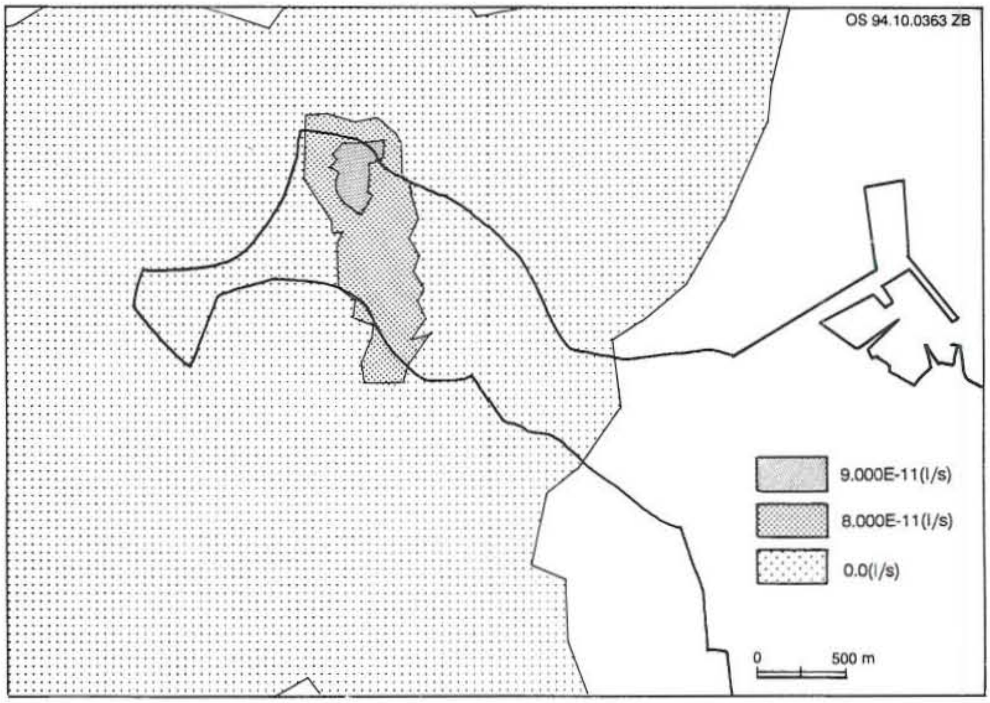

FIGURE 13: Leakage coefficient distribution for the Seltjarnarnes field

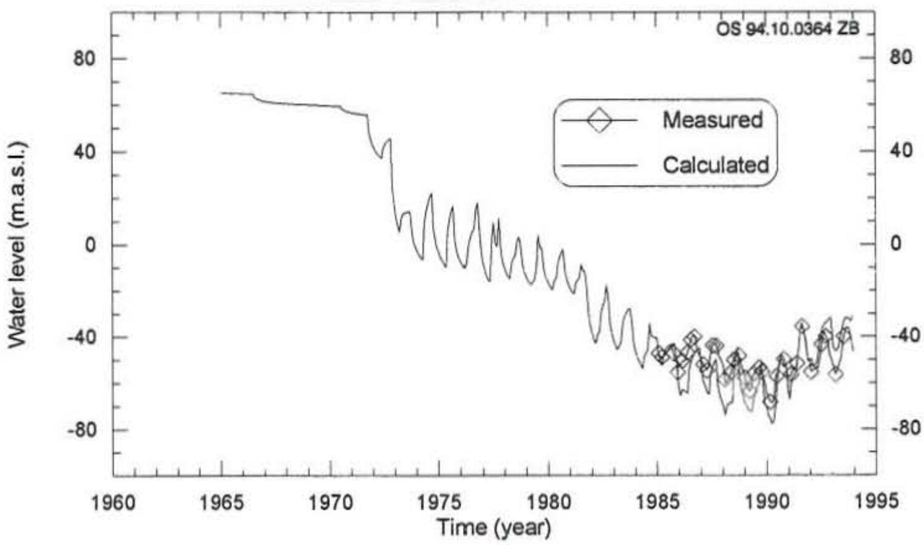

FIGURE 14: Measured and calculated water level in well Sn-01

Future predictions of the water level have been calculated from the year 1994 to 2005. They are done for same two cases discussed before, with and without reinjection:

A. Prediction of future water level without reinjection:

Case 1. The average yearly pumping rate is 35 l/s. It is divided between three wells Sn-4, Sn-05 and Sn-06, with 15 l/s pumped from each well during January till March; 9 l/s during April till September; and 12 l/s during October till December. The results of the calculations are shown in Figure 16. The water level will be at -55 m in the year 2005.

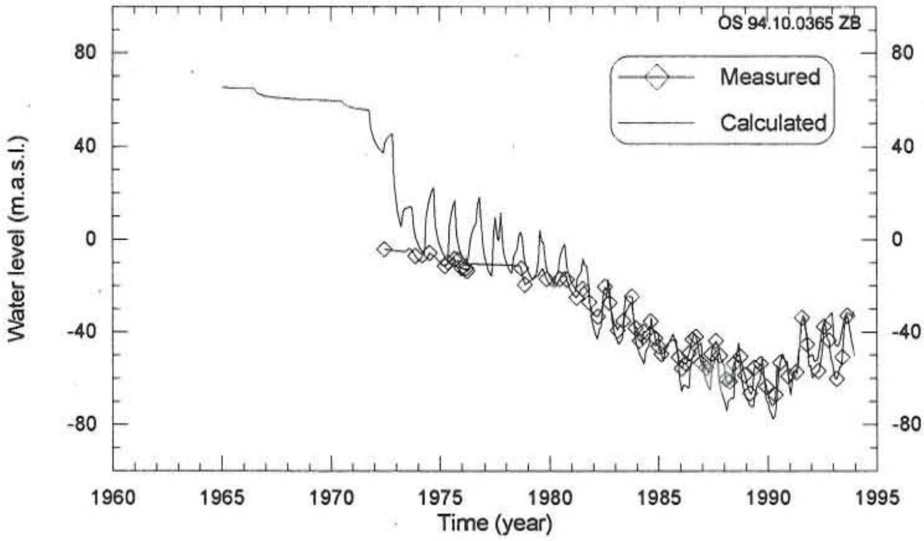

FIGURE 15: Measured and calculated water level in well Sn-02

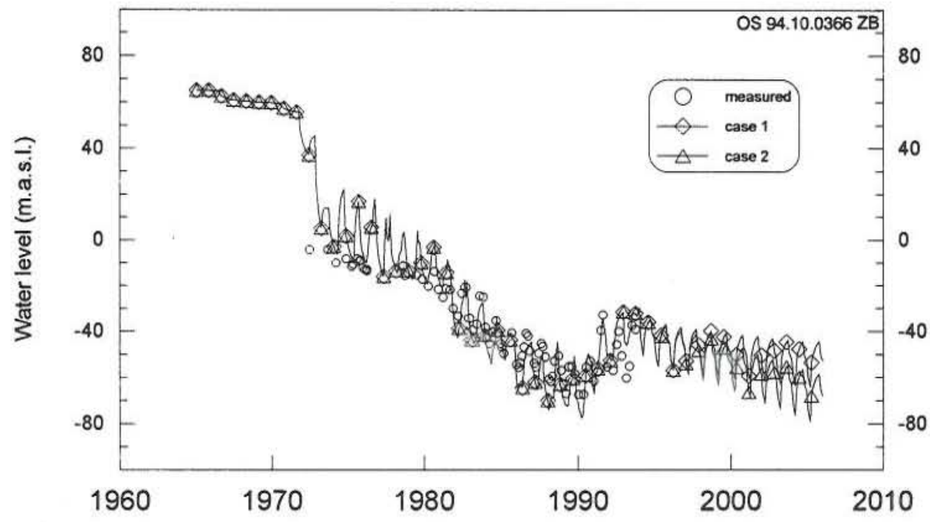

FIGURE 16: Prediction of the water level of the Seltjarnarnes field, without reinjection

Case 2. The pumping rates are the same as in case 1, but with an annual increase of 2%. The results of the calculations are also shown in Figure 16. The water level will be at -70 m in year the 2005.

Comparing the results of the lumped model and distributed model it can be seen that the water level is lower in the lumped model than in the distributed model. The reason is that a closed boundary is used in the lumped model whereas a part of the boundary has a constant water level in the distributed model.

B. Prediction of future water level with reinjection:

Both cases 1 and 2 are considered, with reinjection. The reinjection rate is 20 l/s in well Sn-03. The results are shown in Figure 17. The water level recovers to -5 m in case 1 and to -20 m in case 2.

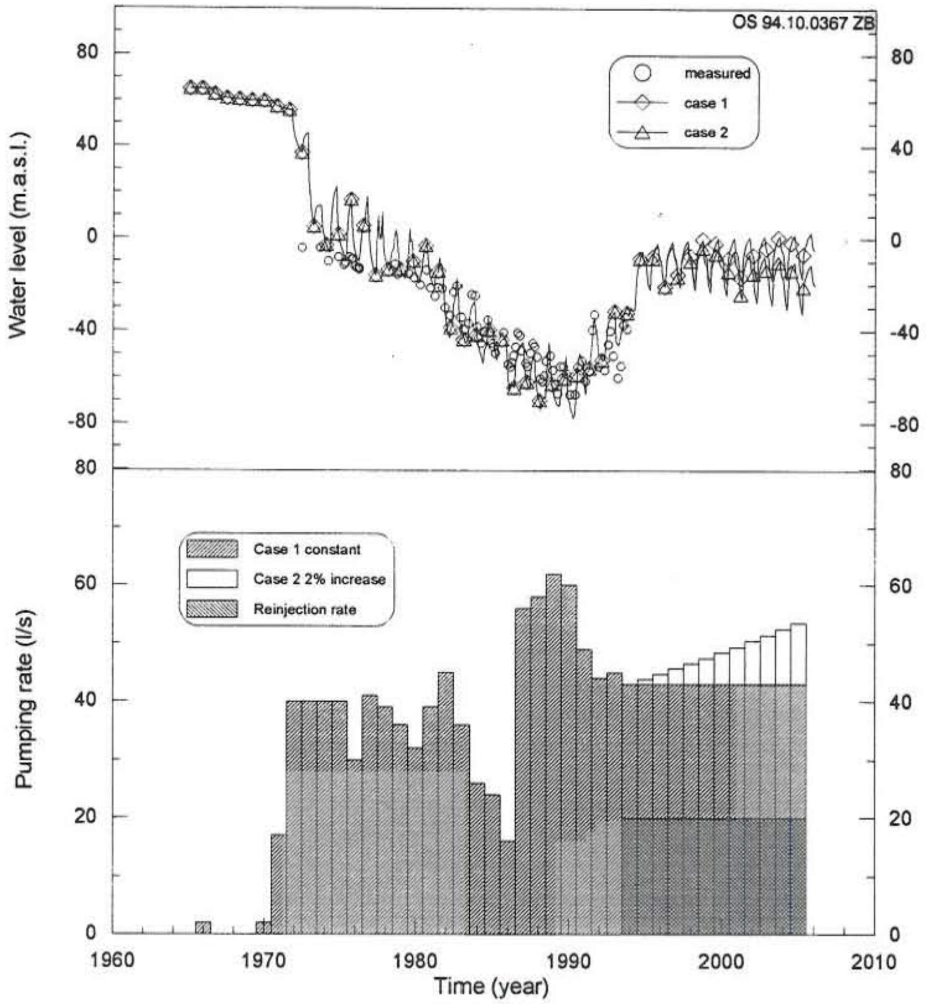

FIGURE 17: Prediction of the water level of the Seltjarnarnes field with reinjection

4.3.2 Mass transport model

Changes of chloride concentration have been observed in the field. It is necessary to simulate the future response during production. A constant initial chloride concentration was assumed in the whole modelling area before production started. The initial value is 500 ppm. The constant chloride concentration boundary condition is 500 ppm. Chloride concentration of the leakage water is 18,000 ppm. The parameters of the model are determined by fitting the observed and calculated chloride concentrations in wells Sn-03, Sn-04 and Sn-05. The results of the calibration are shown in Figures 18, 19 and 20.

Future prediction of the chloride concentration have been performed from the year 1994 to 2005. The two previous cases are considered with and without reinjection. The predictions are also shown in Figures 18, 19 and 20. They show that without reinjection, the chloride concentration increases to 2,000 ppm in the year 2005, but with reinjection the chloride concentration in all three wells decreases and that the difference between cases 1 and 2 is less than 100 ppm. A constant value of 500 ppm is reached in the reinjection well Sn-03. In well Sn-04 a value of 800 ppm is approached and 1250 ppm in Sn-05. So reinjection with low concentration of chloride can prevent the chloride concentration from increasing. The rate of change of chloride concentration is smaller in Sn-05 than in Sn-04 due to different permeability and leakage coefficients in different parts of the area.

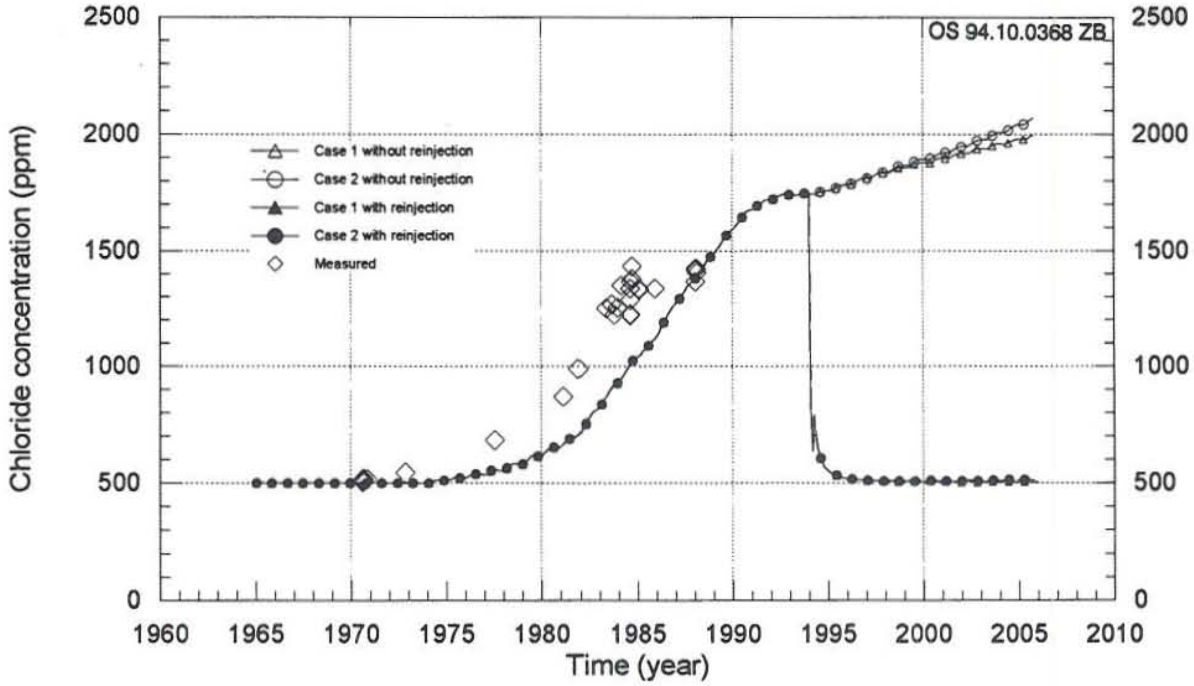

FIGURE 18: Calibration and prediction of the chloride concentration in well Sn-03

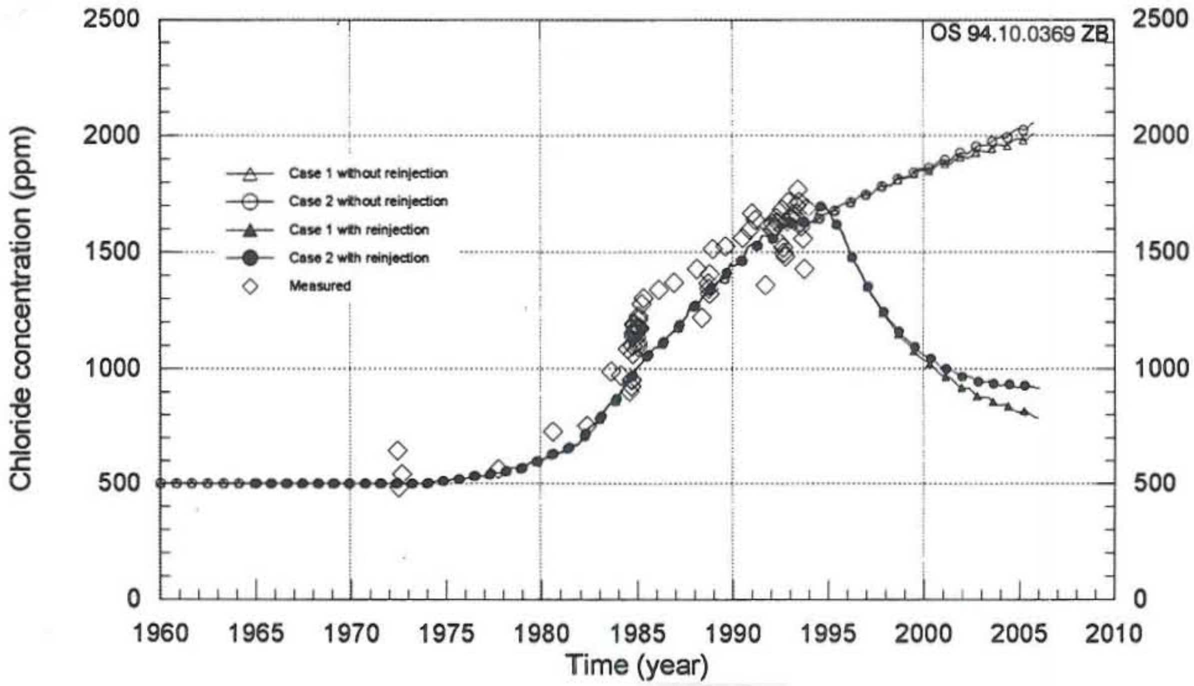

FIGURE 19: Calibration and prediction of chloride concentration in well Sn-04

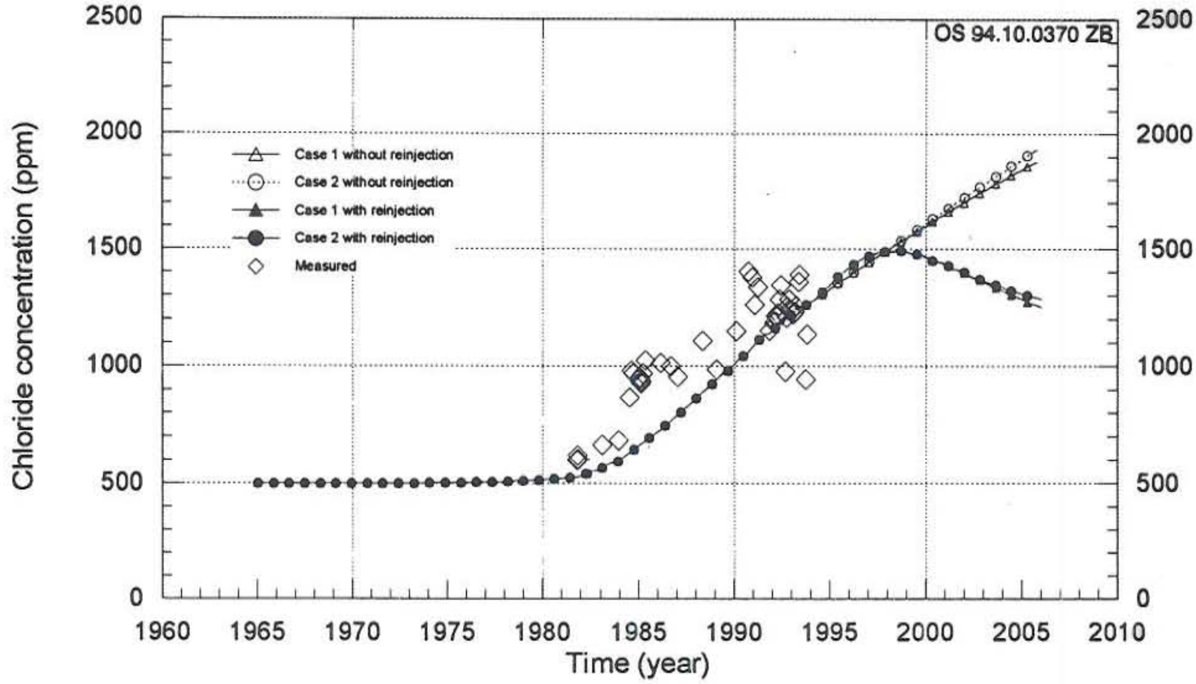

FIGURE 20: Calibration and prediction of chloride content in well Sn-05

4.3.3 Heat transport model

The heat transport equation has been used to calculate temperature changes in the reservoir due to the exploitation. Initial temperature of the reservoir is 110°C. The constant boundary value is 110°C in the southeast part of the area. The temperature of the leakage seawater is 10°C. We can see the effect of the leakage in Figures 21-24. They also show future predictions of temperature that have been made to the year 2005. The effects of production are small or about 0.05°C in 10 years but there is a slight difference between cases with and without reinjection. With reinjection water with the assumed temperature 50°C, the difference in production temperatures in the year 2005 is about 0.2-0.3°C. Because of different permeability and leakage coefficients in different parts of the area, it can also be seen that the rate of change of temperature is smaller in wells Sn-01 and Sn-05 than in Sn-04 and Sn-06.

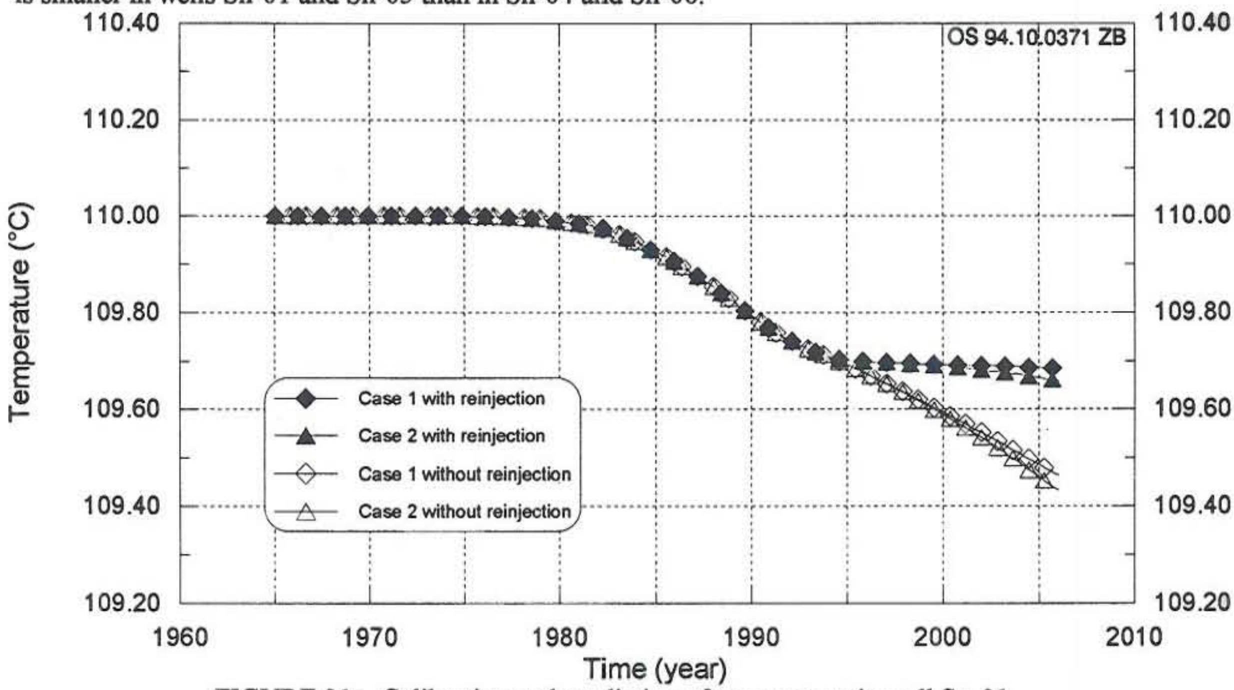

FIGURE 21: Calibration and prediction of temperature in well Sn-01

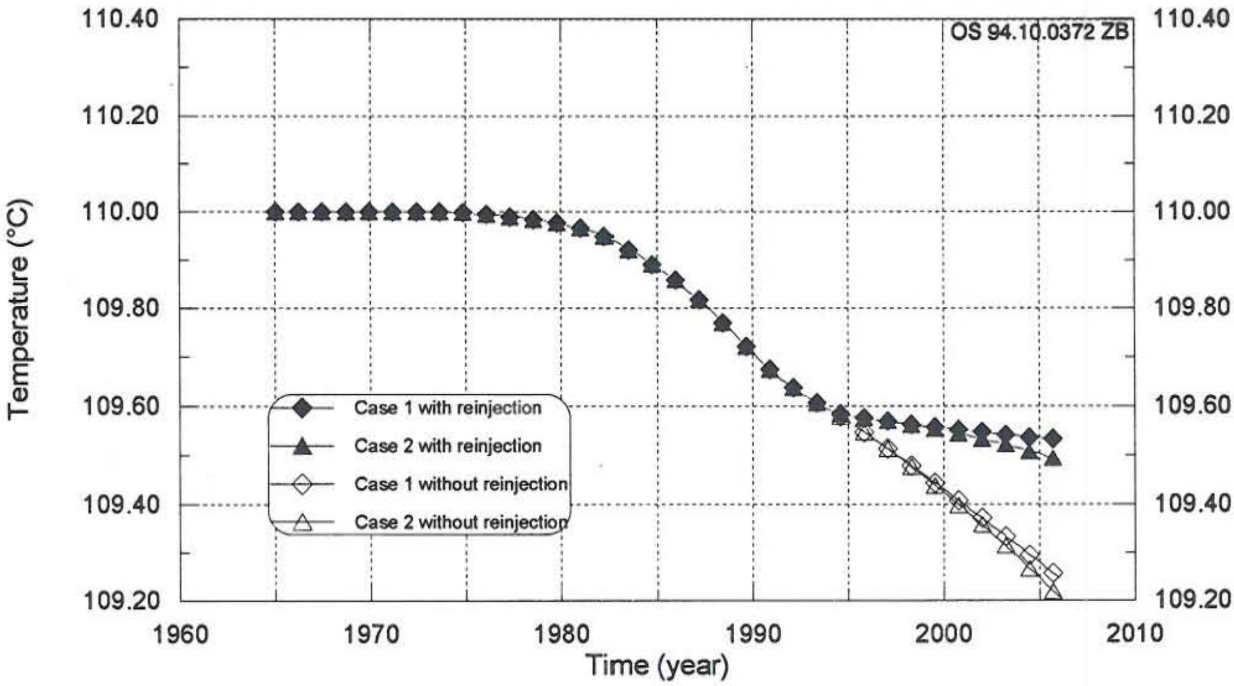

FIGURE 22: Calibration and prediction of temperature in well Sn-04

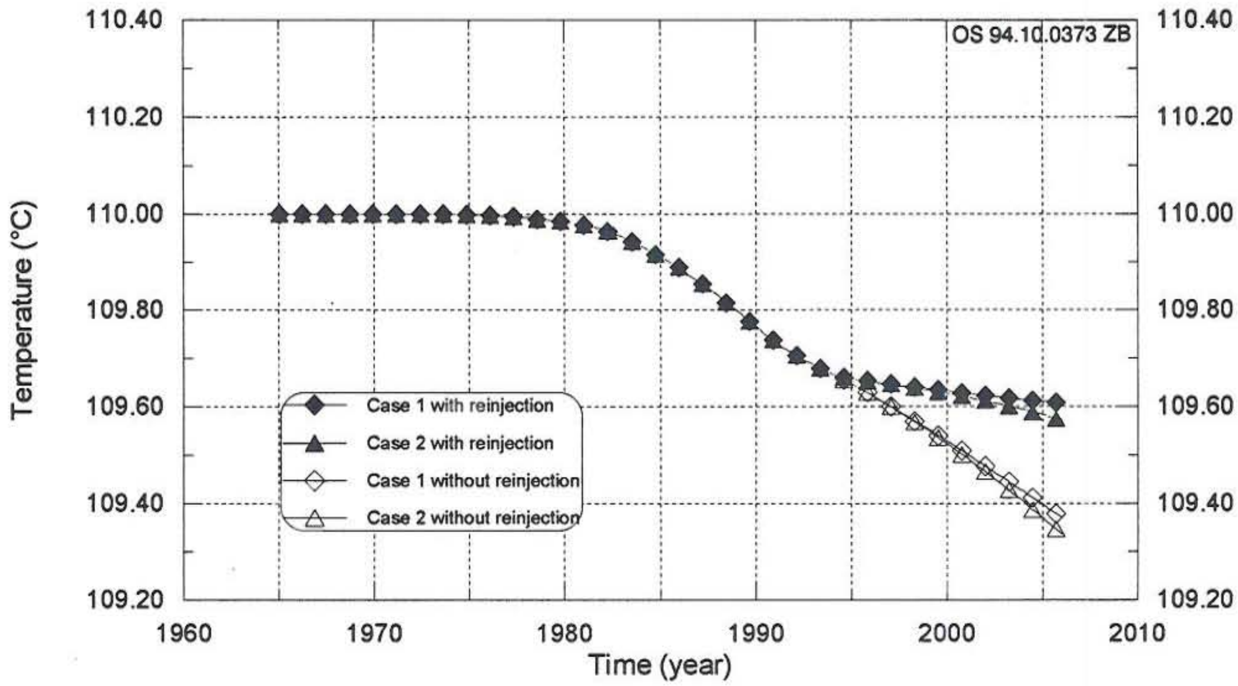

FIGURE 23: Calibration and prediction of temperature in well Sn-05

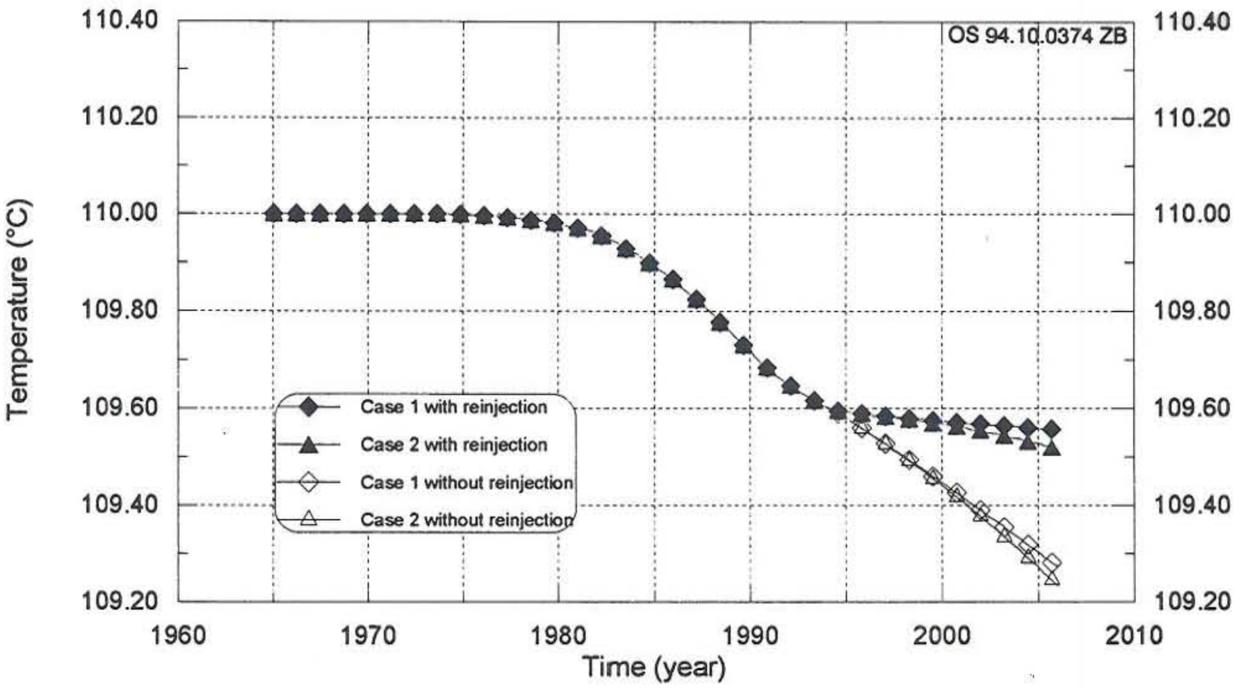

FIGURE 24: Calibration and prediction of temperature in well Sn-06

5. CONCLUSIONS AND RECOMMENDATIONS

1. The geothermal water in the Seltjarnarnes field originates from precipitation falling southeast of the area. Some seawater intrusion has been detected as shown by the rapid increase in chloride concentration.

2. In the Seltjarnarnes geothermal field two kinds of models were used for simulating the fluid flow, chemical mass and heat transport. Lumped models show that the behaviour of the reservoir is controlled by the storativity of the reservoir for a short period of production and by porosity in the upper layer for the long term response.

3. The water level will be at -55 m a.s.l. for an average pumping rate of 35 l/s, but with a 2% annual increase of the pumping rate the water level will be at -70 m a.s.l. in the year 2005. Water level will recover up to -5 m a.s.l. for the first case and to -20 m a.s.l. in the second case, if 20 l/s are reinjected into the reservoir.

4. Reinjection of water with low concentration of chloride and higher temperature than seawater could control the increase of chloride concentration and decrease of temperature.

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